# Mechanisms of Holocene palaeoenvironmental change in the Antarctic Peninsula region

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**ABSTRACT**

**The Antarctic Peninsula is one of the three fastest warming regions on Earth. Here we review Holocene proxy records of marine and terrestrial palaeoclimate in the region, and discuss possible forcing mechanisms for past change, with a specific focus on past warm periods. Our aim is to critically evaluate the mechanisms by which palaeoclimate changes might have occurred, in order to provide a longer-term context for assessing the drivers of recent warming. Two warm events are well recorded in the Holocene palaeoclimate record, namely the early Holocene warm period, and the Mid Holocene Hypsithermal (MHH), whereas there are fewer proxy data for the Medieval Warm Period (MWP) and the Recent Rapid Regional (RRR) warming. We show that the early Holocene warm periods and MHH might be explained by relatively abrupt shifts in position of the Southern Westerlies, superimposed on slower solar insolation changes. A key finding of our synthesis is that the marine and terrestrial records in the AP appear to be show markedly different behaviour during the MHH. This might be partly explained by contrasts in the seasonal insolation forcing between these records. Circumpolar Deep Water (CDW) has been implicated in several of the prominent changes through the Holocene but there are still differences in interpretation of the proxy record that make its influence difficult to assess. Further work is required to investigate contrasts between marine and terrestrial proxy records, east-west contrasts in palaeoclimate, the history of CDW, a long onshore high resolution record of the Holocene, and the role of sea ice in driving or modulating palaeoclimate change, along with further efforts to study the proxy record of the Recent Rapid Regional Warming, and the MWP.**

**Keywords:** Antarctic Peninsula, Southern Ocean, palaeoclimate, Southern Westerlies, ENSO, climate models, Circumpolar Deep Water.

## 1. INTRODUCTION

### Aim and background

The Antarctic Peninsula (AP) is one of the three fastest-warming regions on Earth (3.4 ºC century $^{-1}$, Vaughan *et al*., 2003). The rate of temperature increase is more than five times the global mean (0.6 ± 0.2 ºC during the 20$^{th}$ Century), leading to shifts in species distribution, catastrophic disintegration of ice shelves, retreat and accelerated discharge of continental glaciers (Cook *et al*., 2005; Pritchard and Vaughan, 2007), and the possibility of increased rates of global sea level rise. In order to compare this recent rapid regional (RRR) warming (Vaughan *et al*., 2003) with previous Holocene warm periods and to learn something about the underlying mechanisms, this paper brings together evidence of the mechanisms thought to be driving Holocene climate changes in the AP region. We do this by briefly summarising the temporal and spatial patterns of climate changes in the AP region during the Holocene, with a particular focus on past warm periods. We then discuss the potential forcing mechanisms and provide a critical review that attempts to link these mechanisms to the main climate changes of the Holocene. By concentrating on the Holocene record of warm 'events', the intention is to provide a longer-term context for understanding the RRR warming that is ongoing in the AP today. The final aim of the paper is to identify key gaps in understanding and provide a series of questions that might be addressed by future research and modelling.

This synthesis differs from previous reviews of AP environmental change (e.g., Ingólfsson *et al.,* 1998; 2003; Jones *et al*., 2000; Hjort *et al.,* 2003*;* Domack *et al*., 2003a; Ingólfsson 2004) in that it places greater emphasis on identifying the mechanisms underlying specific periods of climate change.

## Physical setting and location of records

The AP comprises a narrow (<250 km wide) chain of mountains, rising to a maximum of 3500 m (but mostly much lower) along its ~1250 km length (Fig. 1), such that it constitutes a fairly continuous plateau with few gaps below 2000 m. It projects substantially further north than the rest of the Antarctic continent. In many respects the AP is geologically similar to the South American cordillera. It comprises a pre-Jurassic basement overlain and intruded by Jurassic-Tertiary magmatic arc rocks related to eastwards-directed subduction along the western (Pacific) margin of the AP. A comprehensive description of the setting of the Antarctic Peninsula is given by Domack *et al*. (2003b).

### *Terrestrial setting*

The Antarctic Peninsula is heavily glaciated with outlet glaciers terminating in the ocean along both its western and eastern margins. There is a clear boundary, south of which glacier termini feed into ice shelves, and north of which there are no ice shelves (Morris and Vaughan, 2003). This boundary is further south (67-70 °S) on the west side of the AP than on the east (64-68 °S). There are a few extensive ice-free areas where terrestrial palaeoenvironmental records from the Holocene have been preserved, with perhaps the most important being in the north-east Peninsula (James Ross Island, Seymour Island), South Shetland Islands, eastern Alexander Island (Ablation Point Massif), and the coastline around Marguerite Bay. There are numerous small ice-free areas on ridges between outlet glaciers, and on nunataks of the central plateau, but most of these are rocky with few surface deposits or lakes.

### *Atmospheric setting*

In many respects the AP is climatically atypical of the Antarctic continent. It is narrow and so experiences a strong marine influence (Fig. 1), particularly on its western side, which is exposed to the southern westerly winds. The AP is the most northerly part of Antarctica and as such is the most subject to midlatitude influences. In particular, the west side of the AP receives relatively warm, moist air masses derived from mid-latitudes. Moreover, the western AP is the only part of Antarctica where there is a demonstrated correlation between winter temperatures (at Faraday/Vernadsky) and the extent of sea-ice (in the Bellingshausen Sea to the west of the station) (King, 1994) and it has been suggested that sea-ice may play an important role in environmental change around the AP (Vaughan *et al*., 2003). The east side of the Peninsula is less influenced by the Westerlies and experiences a substantially colder and drier climate because of the northwards extension of cold continental air masses from the Antarctic interior into the Weddell Sea embayment (Reynolds, 1981).

*Oceanography*

The Antarctic Circumpolar Current (ACC) flows clockwise in a broad zone around Antarctica, carrying Circumpolar Deep Water (CDW), and is separated from the coast by the katabatic-driven west-flowing Antarctic Coastal Current. The north-south landmass of the Peninsula acts as a barrier to this flow and so the southern boundary of the ACC runs NE along the edge of the AP shelf as the ACC is deflected to the north and through the constriction of the Drake Passage (Fig 1). Thus, the Pacific margin continental shelf of the Peninsula is likely to be affected by any variations in the flow of the ACC. This is particularly important because the ACC propagates changes in oceanic conditions between the Pacific, Atlantic and Indian Oceans (Simmonds, 2003), including the effects of the El Niño Southern Oscillation (ENSO).

Some areas of the continental shelf on the west side of the AP experience intrusions of Circumpolar Deep Water (CDW) (Fig. 1). This is a relatively warm (>1.5 °C), salty (34.65-34.7‰), intermediate depth water mass (Klinck *et al*., 2004) that is derived from modified North Atlantic Deep Water. The top of the CDW is at 200 m depth and so only areas of the shelf deeper than this are affected. These intrusions bring heat and salt onto the shelf. CDW is substantially warmer than typical Antarctic surface waters and so where mixing occurs then intrusions of CDW along the western AP shelf are characterised by surface waters that are above freezing in winter. The inner AP shelf has significant topography and this roughness may contribute to mixing. In places, it is the onshore flow of the ACC that provides the impetus to pump CDW onto the shelf, particularly along the troughs that formed the paths of palaeo ice streams (Smith *et al*., 1999; Smith and Klinck, 2002; Klinck *et al*., 2004).

*Antarctic Peninsula palaeoenvironmental records*

The AP region contains a number of palaeoenvironmental records, each of which provide different information on the patterns, processes and mechanisms of Holocene climate change.

Ice cores provide a record of environmental change in the interior from which it is possible to infer past atmospheric temperatures, average regional sea-ice extent (salts, methanesulphonic acid (MSA)), precipitation, and changes in atmospheric composition (e.g. Aristarain *et al*., 1986; Peel, 1992; Peel and Mulvaney, 1992; Thompson *et al*., 1994).

Lake sediments provide a record of deglaciation, changing biological productivity (linked to temperature), changes in ecology and species composition, and lake ice cover (see review in Hodgson *et al*., 2004). Near-coastal lakes or lagoons can also be used as isolation basins to determine relative sea-level change (Bentley *et al*., 2005a). Epishelf lakes can provide proxy records of ice shelf presence and absence (Bentley *et al*., 2005b; Smith *et al*., 2006, 2007; Roberts *et al*., 2008).

Marine sediment cores from the deep ocean, continental shelf and fjords provide information on changes in sedimentation during deglaciation (e.g. Pope and Anderson, 1992; Pudsey *et al*., 1994; Ó Cofaigh *et al*., 2001; 2005a; Anderson *et al*., 2002; Heroy and Anderson, 2005; Evans *et al*., 2005); palaeoceanographic changes such as surface water productivity (Leventer *et al*., 2002; Sjunneskog and Taylor, 2002; Taylor and Sjunneskog 2002); the proximity and stability of ice shelves and glaciers (Domack *et al*., 1995; 2005; Pudsey and Evans, 2001; Brachfeld *et al*., 2003); duration or extent of sea ice (Leventer *et al*., 1996; Gersonde *et al*., 2003) the influx of meltwater and terrigenous sediments (Domack *et al*., 1994), and clues to the behaviour of ocean currents such as the ACC or distribution of water masses such as CDW (Howe and Pudsey, 1999; Shevenell and Kennett, 2002) or Antarctic Bottom Water (AABW) (Anderson, 1999).

Glacial geomorphological records from terrestrial and continental shelf environments provide additional information on past ice extent and thickness, and the timing of glaciation and deglaciation (e.g. Sugden *et al*., 2006). A variety of onshore features have been used to determine ice sheet extent, including till stratigraphy, moraines, striations, trimlines and erratics (e.g., Clapperton and Sugden, 1982; Rabassa, 1983; Hjort *et al*., 1997; Bentley *et al*., 2006). Offshore, the relatively recent application of high-resolution sonar mapping (i.e., swath bathymetry) to marine glacial geomorphology has allowed identification of landforms such as mega-scale lineations, drumlins, ice- and subglacial meltwater-moulded bedrock as well as moraines and grounding zone wedges, providing detailed information about former ice sheet extent and dynamic behaviour (e.g. Anderson, 1999; Anderson *et al*., 2002; Canals *et al*., 2000; 2002; Ó Cofaigh *et al*., 2005a; 2005b; Evans *et al*., 2005; Wellner *et al*., 2006).

Collectively these historical records provide valuable palaeoenvironmental data on the patterns of, and mechanisms for, Holocene environmental change on the AP. However, there is still a relative paucity of high-resolution, long (full Holocene) climatic records. This has made it difficult to assess the regional significance of some climatic events during the Holocene and to compare relative forcing factors. The main patterns of Holocene climate change have to therefore be pieced together from a compilation of the marine, continental shelf, and terrestrial records.

In the next section we provide a brief synthesis of Holocene climate change in the AP region based on information from lake records, marine records, and compilations of Antarctic ice core records. Figure 1 shows the location of all sites discussed in this synthesis.

## 2. THE PATTERNS OF AP CLIMATE CHANGE DURING THE HOLOCENE

We discuss below the various time periods in which proxy records demonstrate a significant warming, namely the early Holocene, mid-Holocene warming, Medieval Warm Period, and the proxy and instrumental record of RRR warming. We give all ages in calibrated years before present (cal yr BP) unless stated otherwise. In those cases where we have calibrated published $^{14}C$ dates we used CALIB5.0 (Stuiver and Reimer, 1993), and for marine material we utilised a marine reservoir correction of 1300 ± 100 yr (i.e. $\Delta R = 900 \pm 100$ yr) (Berkman and Forman, 1996). We do not discuss here the problems with radiocarbon dating of Antarctic marine sediment as this has been reviewed elsewhere (e.g. Anderson et al., 2002; Ohkouchi and Eglinton, 2008).

### Early Holocene climate optimum (c. 11-9.5 cal ka)

A growing number of proxy records demonstrate that there was a period of significant warmth in Antarctica during the early Holocene. For example, syntheses of stable isotope proxy records of temperature from ice cores around Antarctica show a widespread early Holocene climatic optimum ca. 11,000-9,500 cal yr BP (Ciais *et al*., 1992; Masson *et al*., 2000; Masson-Delmotte *et al*., 2004) (Fig. 2a).

Although ice sheet retreat around the AP may have begun as early as 18.5 cal ka BP, in most areas retreat was well underway by 14-13 cal ka BP (Heroy and Anderson, 2005; Evans *et al*., 2005), and continued through the early Holocene warmth. For example, Pudsey *et al*. (1994) suggested a deglaciation age between ~ 13 and 12 cal ka BP from the continental shelf off Anvers Island and the onset of glaciomarine sediments overlying diamicton in the Palmer Deep record (Fig. 2b) show deglaciation and an increase in primary production and iceberg rafting at c. 11-10 cal ka BP (Domack, et al., 2001; Domack, 2002; Leventer *et al*., 2002). In some areas rapid deglaciation reached onto the inner shelf and fjords (Harden *et al*., 1992; Pudsey *et al*., 1994; Shevenell *et al*., 1996; Evans *et al*., 2005). During this period there may have been retreat of ice stream grounding lines landward across the continental shelf (Ó Cofaigh *et al*., 2005a). The exquisite preservation of landforms and lack of sequential retreat moraines on the continental shelf in Marguerite Bay suggests that the ice streams probably thinned and retreated rapidly in this area. In contrast, on the east side of the AP, grounding-zone wedges indicate that ice sheet retreat was more gradual, punctuated by still-stands across shallower shelf regions (Evans *et al*., 2005).

At the same time as circum-Antarctic ice cores record the early Holocene optimum the Palmer Deep record is characterized by an apparent 'cold' proxy record: lower diatom abundance and an assemblage characteristic of more persistent sea ice (Taylor and Sjunneskog, 2002; Sjunneskog and Taylor, 2002), higher coarse-fraction (gravel) abundance, higher magnetic susceptibility and mass accumulation rates indicating greater terrigenous input from 11.5 – 9.07 cal ka BP (Domack, 2002) (Fig. 2b). Similarly, the benthic foraminiferal isotope record shows that cold shelf water occupied the site immediately following glaciation and persisted into the early Holocene (Ishman and Sperling, 2002).

This period of early Holocene warmth is therefore evident in Antarctic records from ice cores, some marine cores and geomorphological records of continuing deglaciation. In contrast, the Palmer Deep record implies relatively cold conditions during this time. Furthermore, the timing of deglaciation varied on either side of the AP with some evidence of earlier deglaciation on the west side than on the east (Ó Cofaigh *et al*., 2005a; Evans *et al*., 2005; Sugden *et al*., 2006; Hodgson *et al*., 2006).

*After the optimum (9.5-4.5 cal ka BP)*

The period between the early Holocene optimum and Mid-Holocene warmth shows complex patterns of change in different parts of the AP. For example, immediately post-dating the early Holocene climate optimum there was a retreat of George VI Ice Shelf on the west side of the AP, with the onset of collapse at 9595 cal yr BP and complete or partial re-formation by 7945 cal yr BP (Bentley *et al*., 2005b; Smith *et al*., 2007, Roberts *et al.* 2008). This coincides with evidence of the southward intrusion of warmer, more subpolar waters at 9000-6700 cal yr BP in the Palmer Deep (Leventer *et al*., 2002). However, on the eastern side of the AP the Larsen Ice Shelf - B (LIS-B) remained intact (Domack *et al*., 2005) throughout, but it is not clear (Hodgson *et al*., 2006) if this is because (i) early Holocene temperatures did not rise sufficiently at this location to trigger collapse, or (ii) warming did reach LIS-B, but it was too thick to be affected by the meltwater facture mechanism believed to cause ice shelf collapse (Domack *et al*., 2005).

Deglaciation was ongoing during this period but possibly at a slower rate as the grounding line moved onto the inner shelf. Sedimentation commenced in newly exposed lake basins in the north-eastern AP and some islands to the north (Ingólfsson *et al*., 1998; 2003, Jones *et al*., 2000). On the west side of the AP, significant glacier thinning and ice margin retreat continued until at least 7-8 cal ka BP (Bentley *et al*., 2006). Parts of the coast on King George Island in the South Shetland Islands, were ice-free by ~ 9.5 cal ka BP and some lake basins began to accumulate sediments ~ 9.5-9.0 cal ka BP (Mäusbacher *et al*., 1989; Schmidt *et al*., 1990; Hjort *et al*., 2003; Bentley *et al*., 2005a), but other areas in the same island chain did not become free of ice until much later in the Holocene. For example, parts of Byers Peninsula on Livingston Island seem to have remained ice-covered until as late as 5-3 cal ka BP (Björck *et al*., 1996a).

Further north, marine cores from the South Atlantic (50-53 deg S) suggest the onset of cooling between 9 and 7 cal ka BP (Bianchi and Gersonde, 2004), and sea-ice expansion and surface ocean cooling after 9.3 cal ka BP (Nielsen *et al*., 2004).

**Mid-Holocene warm period (4.5-2.8 cal ka BP)**

It was not until the mid-Holocene that the next period of significant warmth occurred in the AP. This interval is reviewed in detail in Hodgson *et al*. (2004). The best-dated records place it between either 4000 to 2700 $^{14}$C yr BP (4500 to 2800 cal yr BP) in the Antarctic Peninsula region (Björck *et al*., 1991a) or 3800 to 1400 cal yr BP just to the north of the AP (Hodgson and Convey, 2005; Jones *et al*., 2000) (Fig. 2). This Mid-Holocene Hypsithermal (MHH) is detected as a period of rapid sedimentation, high organic productivity and increased species diversity in lake sediments ranging from the South Shetland Islands (Schmidt *et al*., 1990; Björck *et al*., 1996a) and James Ross Island (Björck *et al*., 1996b) to maritime Antarctic islands such as Signy Island (Jones *et al*., 2000; Hodgson and Convey, 2005) and subantarctic South Georgia (4400-2400 cal yr BP; Rosqvist and Schuber, 2003) (Fig. 2c, e). In marine sediments it is detected as a period of reduced sea ice coverage, and greater primary production, along with an increase in meltwater-derived sedimentation. For example, it has been detected in multi-proxy analyses from Lallemand Fjord, western AP (Fig. 2d) (Shevenell *et al*., 1996, Taylor *et al*., 2001, Domack *et al*., 2003b) and has also been associated with collapse of the Prince Gustav Channel Ice Shelf in the northern Peninsula between *c*. 5000 and 2000 cal yr BP (Pudsey and Evans, 2001), and fluctuations of the Larsen-A Ice Shelf between 4000 and 1400 cal yr BP (Brachfeld, *et al*., 2003; Pudsey *et al*., 2006). Terrestrial sediments, such as moss banks (Björck *et al*., 1991b) also show evidence of milder climate in the interval 4150-1840 $^{14}$C yr BP (~ 4700 to 1800 cal yr BP). Sites in the northern AP show increased amounts of South American pollen in lake sediments during this period (Björck *et al*., 1993).

There is some ice core evidence for a mid-Holocene warm period. For example, the 4000-year long Plateau Remote record (Mosley-Thompson, 1996) shows that c. 4 to 2.5 cal ka BP was substantially warmer than the most recent 2.5 kyr and may have included two particularly warm periods, centred on 3.6 and 2.8 cal ka BP. Ciais *et al*. (1994) synthesised a series of ice core records and demonstrated that there was some evidence of relative warmth from 4.5 to 2 cal ka BP. However, not all ice cores show significant warming during this period. Similarly, the warming is not seen in all marine records, a notable example being the Palmer Deep record where a period of ocean warmth, based on increased biological production and decreased concentrations of ice rafted debris, started at 9000 cal yr BP and continued until c. 3600 cal yr BP (Domack, 2002) (Fig. 2b). In other words, in the Palmer Deep record there is evidence for the continuous presence of relatively warm UCDW on the continental shelf for a period of over 5000 years with no obvious cooling between the onset of the early Holocene warmth and the MHH. Moreover, in the Palmer Deep record relatively warm water disappears abruptly at 3600 cal yr BP, whereas in the lake records the MHH tails off more slowly.

In summary, whilst there is widespread agreement on the presence of some sort of warm period in the mid-Holocene, the precise timing of the event often varies by hundreds of years, either because the timing varied spatially, or because there are insufficient numbers of dates (Kulbe *et al*., 2001). When comparing marine with terrestrial lake records, part of the disparity may be the result of inherent difficulties in dating Antarctic marine sediments such as uncertainties in the Antarctic marine reservoir effect, and the influence of reworked organic carbon. Moreover, there is no clear presence of a MHH in the Palmer Deep proxy records, creating a substantial challenge for coherent explanations of the mechanism for mid Holocene warmth.

*After the optimum (2.5-1.2 ka): a Neoglacial interval ?*

The end of the MHH has been suggested by Kulbe *et al*. (2001) to have been marked by a pronounced shift to colder climate conditions recorded in both the Vostok and Komsomolskaya ice cores after 2500 cal yr BP. At the same time glaciers readvanced into Lallemand Fjord (Domack and McClennen 1996). As noted above, in the Palmer Deep marine record the onset of this neoglacial was earlier, at 3.6 cal ka BP, as shown by a decrease in Mass Accumulation Rate (MAR) and increase in coarse-fraction IRD (Domack, 2002). In general, diatom abundance and assemblages are consistent with alternating periods of more intense (perennial) sea-ice and open water and surface waters that were never warm for long enough for subpolar species to become established (Taylor and Sjunneskog, 2002; Sjunneskog and Taylor, 2002). Evidence of glacier advance on the Peninsula during the Neoglacial is not yet well-constrained (Domack, 2002): numerous studies have identified Late Holocene glacier advances but most are poorly dated or even undated, and some of the putative Neoglacial advances may belong to a Little Ice Age (see Ingólfsson *et al*., 1998 for review). However, there is good evidence that the Prince Gustav Channel Ice Shelf started to reform after 1900 $^{14}C$ yr BP (date corrected for reservoir effect and core top age but not calibrated) (Pudsey and Evans, 2001; Pudsey *et al*., 2006) and the Larsen-A Ice Shelf reformed by 1400 cal yr BP (Brachfeld *et al*., 2003) as the climate began to cool: numerous biological proxy records in lakes and other sites show a temperature-related decline in production at about this time (Björck *et al*., 1991a; Jones *et al*., 2000, Hodgson and Convey, 2005).

**Medieval warm period (1.2–0.6 cal ka BP)**

A period of warmer climate, termed the Medieval Warm Period (MWP), has been identified between about c. 800 and 1400 AD (~1200 to 600 cal yr BP) (some studies use 800-1200 AD e.g., Broecker,

2001) in many Northern Hemisphere records. However, its existence in Antarctica, or even the Southern Hemisphere has not been unequivocally established from proxies capable of sub-decadal resolution (Mann and Jones, 2003; Broecker, 2001). In the AP, records of an event equivalent in timing to the MWP are restricted to those obtained from marine cores. For example, Domack *et al*., (2003b) interpreted the record from Lallemand Fjord as showing a lesser TOC maxima following the Mid-Holocene TOC peak, and that it may correspond to increased productivity during the MWP (Fig. 2d). Domack *et al*. (2003b) also reported a MWP signal from a short core in the Andvord drift, also ending at about 700 cal yr BP. Khim *et al*. (2002) analysed a marine core close to the western AP, and showed higher magnetic susceptibility values in the interval 1250-1450 AD (~ 750-550 cal yr BP), which they interpreted as a signal of warmer surface water temperatures. In summary, there is some evidence of a warm event occurring in the AP at approximately the same time as the MWP, but so far the proxy records are mostly limited to marine records.

*Little Ice Age*

Like the MWP, evidence for an LIA in AP proxy records is patchy but it is recognized in the Palmer Deep for the period 700 to 150 cal yr BP with more persistent sea ice and colder sea-surface and bottom-water conditions corresponding with local glacial advances (Domack *et al*., 1995; 2003b; Shevenell *et al*., 1996; Leventer *et al*., 1996, 2002; Shevenell and Kennett, 2002; Taylor and Sjunneskog, 2002; Sjunneskog and Taylor, 2002; Warner and Domack, 2002). Various outlet glaciers or ice shelves such as Rotch Dome, Livingston Island (Björck *et al*., 1996a) and the Müller Ice Shelf (Domack *et al*., 1995) are thought to have advanced during this time. However, the precise timing of those advances is well-constrained at only a few sites, and many of the terrestrial records of glacier advances are as yet undated. There is very limited evidence of a LIA from lake proxy evidence. For example, Liu *et al*. (2005) show a decline in penguin populations on Ardley Island, South Shetland Islands between 450-200 cal yr BP based on their study of ornithogenic lake sediments.

**Recent Rapid Regional warming**

Instrumental measurements show the spatial pattern and magnitude of the RRR, and in particular the pronounced contrast between west (more warming) and east (less warming) sides of the AP. In proxy records, the RRR warming is seen in increased sediment accumulation rates in some northern maritime AP lake cores (Appleby *et al*., 1995), and some high-resolution marine cores. For example, Domack *et al*. (2003b) described a core from the Gerlache Strait that showed increases in % TOC and relative abundance of warm water species of diatoms in the upper 15 cm (c. 30 yr) of the record followed by increases in IRD, and % clay, presumed to be from increased meltwater (Domack *et al*., 2003b). The laminated marine sediments found in the mid-Holocene portions of some cores are not yet seen in any recent marine sediments, leading Domack *et al*. (2003b) to suggest that despite the RRR warming, conditions have not yet reached those that occurred *c*. 4000 yr BP along the west side of the Peninsula.

The Gomez and Dolleman Island ice core records show the RRR warming (Peel *et al*., 1988) but other AP ice cores do not show up recent warming in their isotopic proxies. For example, the James Ross Island core (Aristarain *et al*., 1986) shows isotopic cooling in the last century (Mosley-Thompson and Thompson, 2003). It may be that several of the ice cores are simply not in areas representative of where the strongest warming is happening now. In other words, the records show a weak spatial coherence with the locations of modern climate change (King and Comiso, 2003).

In summary the RRR warming is recorded in some proxies, but few studies have yet focussed on this period in the proxy records at sufficiently high resolution.

## 3. POTENTIAL FORCING MECHANISMS

The Holocene climate changes that have occurred in the AP region, outlined above, have been brought about by a number of forcing mechanisms operating at different relative strengths at different times. These potentially include the long-term influence of changes in orbital solar forcing and greenhouse gases superimposed on shorter term changes in the configuration of heat transport within the ocean and atmosphere. We briefly list the main potential forcing mechanisms below.

### Greenhouse gases

Figure 3b shows the concentrations of two greenhouse gas concentrations determined from ice core analysis (Raynaud *et al*., 2000), smoothed to yield the long-term Holocene changes (Renssen *et al*., 2005). The concentration of $CO_2$ increased from ~260 ppm to ~280 ppm at pre-industrial times (Fig. 3b). Since then it has increased to over 380 ppm. Methane content of the atmosphere first decreased from ~660 ppb at 9 cal ka BP to ~ 580 ppb at 5 cal ka BP, and then increased to its pre-industrial value of ~710ppb (Fig. 3b), followed by rapid increase to over 1800 ppb.

### Solar forcing

Variations in solar insolation brought about by the Milankovitch cycles can influence the AP region either as a direct result of changing insolation over the region itself, or indirectly through changes in another region (e.g., the northern high latitudes) bringing about a change in global ocean circulation that then propagates to the AP. Orbital calculations (Berger and Loutre, 1991) show annual insolation at the latitude of the central AP (65°S) has been decreasing over much of the Holocene (Fig. 3a). The insolation values show marked seasonal contrasts. For example, although the sum of *annual* insolation received at

65°S rose to a peak in the mid-Holocene and has declined since, the amount of insolation received during summer (January) has actually increased steadily since the early Holocene. Not all seasons have shown a simple pattern of decrease or increase through the Holocene: for example, the insolation in the months October to January reached their Holocene maxima progressively later (Fig. 3a).

The potential importance of insolation and greenhouse changes was demonstrated by Renssen *et al.* (2005) who carried-out a model study of Holocene climate evolution of Antarctica and the Southern Ocean. This Antarctic climate simulation relied on a coupled atmosphere-sea-ice-ocean-vegetation model, run over the global domain, but with a specific focus on the climate of the last 9 kyr in the Antarctic. The model was forced with orbitally-driven variations in insolation, and ice-core-derived changes in greenhouse gas ($CO_2$ and $CH_4$) concentrations (Raynaud *et al.*, 2000). Other forcings (solar constant, other gases, ice-sheet configuration) were held fixed at pre-industrial (1750 AD) values.

Because of the complex seasonal/monthly patterns of insolation change through the Holocene (Fig. 3a), the resulting simulated temperature variations show a strong dependence on the month or season in question. Springtime (October) temperatures show a steady decline from the earliest parts of the simulation (9 ka) to present (Fig. 3d). However, in summer (January) simulated temperatures are slightly higher (+0.5ºC) than present in the early Holocene, rising to a maximum, or optimum (+1.3ºC warmer than present), in the mid-Holocene followed by a relatively rapid cooling to present-day values (Fig. 3d). Autumn (April) temperatures show an increase in temperature throughout the duration of the experiment. Winter (July) temperatures decrease throughout the Holocene but with the majority of cooling occurring between 9 and 5 cal ka BP. These results are spatially-averaged for the Antarctic as a whole but Renssen *et al*. (2005) also discussed the spatial pattern of change at key intervals. Specifically the relative warmth in early Holocene summer is concentrated in West Antarctica and to the west of the Antarctic Peninsula, whereas the coast of East Antarctica shows cooler than present conditions. Similarly, the warmth associated with the mid-Holocene summer optimum is concentrated over West Antarctica but is also seen to a lesser extent around East Antarctica.

Other features of the model simulations include a difference in timing of the thermal optimum between West (earlier optimum) and East Antarctica (later optimum) during the early Holocene. This effect was attributed to a more reduced sea-ice cover in the West compared to the East (Renssen *et al*., 2005). The model shows that the Westerlies appear to have been stronger in winter (July) at 9 cal ka BP than present, but all other seasons show a gradual increase in wind strength through the Holocene. The ACC showed strongest flow at 9 cal ka BP, weakened up to 5 cal ka BP, and then increased again, but the magnitude of these changes in flow rate are relatively small (~3%).

It is clear from the model simulation of Renssen *et al*. (2005) that the effect of relatively simple orbital and greenhouse gas forcing is a complex pattern of environmental change in Antarctica. We suggest here that the model simulations also have other clear implications for the understanding of AP palaeoenvironmental change. Firstly, proxies that reflect processes operating at different times of the year may have reached optima at different times during the Holocene. For example, proxy records of biological productivity in lakes (summer bloom) may have responded out of phase with some records of productivity in the marine realm (spring bloom). Secondly, proxy records that reflect a specific season, such as summer productivity, may yield a different signal to those proxies that record more of an annual 'average' (e.g., some ice core records). These differences apply to both biological and physical proxies. This is analogous to the recent work on ice shelf collapse which has shown the importance of peak summer temperatures for determining ice shelf stability (Vaughan and Doake, 1996; Scambos *et al*., 2000). Therefore, past summer, rather than annual temperatures will be one of the important factors when trying to determine drivers for past Holocene ice shelf collapse.

**Ocean circulation**

Some authors have attributed Holocene changes in the Southern Ocean to Northern Hemisphere thermohaline change (Hodell *et al*., 2001; Nielsen *et al*., 2004). Alternatively, a number of authors have suggested that many of the changes in the AP during the Holocene are too rapid to be explained by changes in Northern Hemisphere oceanography (Shevenell and Kennett, 2002; Ishman and Sperling, 2002), or that such explanations are unnecessary to explain Antarctic palaeoclimate (Renssen *et al*., 2005). Instead, oceanographic changes in the ACC and South-east Pacific, described below, have been invoked as oceanographic drivers of AP climate change.

*The role of the Antarctic Circumpolar Current*

The southernmost edge of the ACC abuts the AP shelf as it is funnelled through the Drake Passage and eventually into the Atlantic sector of the Southern Ocean (Fig 1). By impinging on the AP the ACC is thought to play a fundamental role in the climate of the west coast of the AP and is also associated with the upwelling of warm CDW (Smith *et al*., 1999; Smith and Klinck 2002). The northern part of the ACC transports sub-antarctic water to the Chilean coast between 40 and 45° S, where it splits into the poleward Cape Horn current, and the equatorward Peru Chile Current (PCC). Therefore, the ACC delivers relatively cold nutrient rich waters to the southeastern Pacific via the PCC, and consequently it has been argued that the Holocene evolution of the PCC has been controlled by latitudinal shifts of the Westerlies-driven ACC (Lamy *et al*., 2001; 2002). Because of this advection of ACC-derived water, changes in the PCC may provide a proxy for behaviour of the ACC (Lamy *et al*., 2001; 2002).

Lamy *et al*. (2001; 2002) reconstructed the PCC over the last 8000 yr based on a multi-proxy approach including sea-water palaeotemperature (Fig. 2d). They suggested that higher palaeotemperatures in the PCC between 7000 and 4000 cal yr BP most likely reflect a decreased advection of cold and nutrient rich water by the ACC and attribute these changes to a poleward displacement of the Southern Westerly wind belt.

Other studies have also highlighted the link between the South Pacific and the AP (Shevenell and Kennett, 2002; Ishman and Sperling, 2002; Smith *et al*., 2007). For example, it has been argued that variability in the flux of Upper Circumpolar Deep Water (UCDW) on the western AP continental shelf is linked to atmospheric and oceanographic circulation in the South Pacific (Shevenell and Kennett, 2002; Ishman and Sperling, 2002). Specifically, Southern Ocean changes may result from low- to high-latitude atmospheric teleconnections involving Southern Hemisphere westerly wind field fluctuations driven by changes in the South Pacific (Klinck and Smith, 1993; Charles *et al*., 1996; Ninnemann *et al*., 1999; Lamy *et al*., 2002; Shevenell and Kennett, 2002). Specifically, Shevenell and Kennett, (2002) used their foraminiferal-based isotope study from the Palmer Deep to propose that Holocene changes in CDW were controlled by the strength or position of the Southern Hemisphere westerly wind field.

**The Southern Westerlies**

The western AP just reaches the southern boundary of the Southern Hemisphere westerly winds (e.g. Trenberth, 1987) and as such is sensitively positioned to monitor southward shifts in this wind system. The position of the Southern Westerlies is dependent on steep sea surface temperature (SST) gradients within the ACC but also to the location of the south-east Pacific anticyclone in the north, and the circum-Antarctic low pressure belt in the south (Pittock, 1978; Aceituno *et al*., 1993). The westerly wind belt can therefore be deflected southwards when there is a strong 'blocking' South-east Pacific anticyclone, and deflected north by cooling and increased sea-ice around Antarctica.

Toggweiler *et al*. (2006) show from observational and model data that there is a general relationship between the position of the westerlies and climate: warm climates like the present tend to have poleward-shifted westerlies; cold climates (e.g., the LGM) have equatorward-shifted westerlies. Recent observations back this up with evidence of a poleward shift of the Westerlies during the last 40 years, which models suggest is a response to anthropogenic warming (Shindell and Schmidt, 2004). More recently, Marshall *et al*. (2006) concluded that the RRR warming in the northern Peninsula has been due to increased westerly wind strength, driven by increased greenhouse gas concentrations. Thus, it is possible that poleward displacements of the Southern Westerlies could be involved in past periods of higher temperature and precipitation on the AP.

**The El Niño Southern Oscillation (ENSO)**

Recent measurements show that on sub-decadal timescales the ENSO can have a profound effect on near-coastal hydrography in the AP. For example, Meredith *et al*. (2004) report the impacts of the 1997/98 ENSO event on the oceanography and climate of the Marguerite Bay area, western AP. Year-round hydrographic casts and meteorological observations showed that the winter of 1998 was characterised by low sea-ice concentrations, high atmospheric temperatures and a high frequency of northerly winds. Thus, if the Meredith *et al*. (2004) findings are representative of the western AP then the winter following an El Niño year appears to be associated with relative warmth in this region. Similarly, Harangozo (2000) showed from recent meteorological data that ENSO affects the AP, and in particular the Amundsen-Bellingshausen Sea and Weddell Sea areas. During El Niño events the Amundsen-Bellingshausen Sea region shows anomalously high pressures whilst the Weddell Sea shows somewhat lower pressures.

The corollary of this teleconnection is that (palaeo)climate changes in the AP region during the Holocene may be related to changes in the intensity or frequency of ENSO (Simmonds, 2003). There are a growing number of proxy records of the Holocene evolution of ENSO that we can use to evaluate this forcing mechanism. For example, the frequency of clastic laminae in a lake record in southwestern Ecuador has been linked to suppression of El Niño prior to 7000 cal yr BP (Moy *et al*., 2002) (Fig. 3c). Past ENSO variability has also been inferred from mollusc analysis at archaeological sites located on the north and central coast of Peru (Sandweiss *et al*., 2001), and from palaeolimnological studies in the Galapagos Islands (Riedinger *et al*., 2002). These suggest that after 7000 cal yr BP ENSO frequency increased, peaking at 1200 cal yr BP (Rodbell *et al*., 1999; Moy *et al*., 2002). Numerical experiments using a coupled ocean/atmosphere model have shown that seasonal insolation due to changes in the Earth’s orbital parameters might explain the suppression of El Niño before 7000 cal yr BP and the increasing ENSO variability after that date (Clement *et al*., 1999, 2000). Consequently, if ENSO has played a dominant role in forcing AP climate then winter air temperatures along the western AP should have started to warm (and sea-ice coverage decreased) around 7000 cal yr BP, and should have shown significant variability since then, reaching maximum warmth (sea ice reaching a minimum extent) at 1200 cal yr BP.

**Sea-ice interactions and feedbacks**

The western Antarctic Peninsula is the only region of Antarctica where a clear relationship has been identified between historical (winter) temperature and sea-ice extent (King, 1994; King and Turner, 1997; Jacobs and Comiso, 1997). This link is crucial to the climate of this region as it provides a potential positive feedback that can amplify climate change from atmospheric or oceanic causes (Vaughan *et al*., 2003). Thus, it has been suggested that the role of sea-ice in the AP has probably not been so much a primary driver as it has been a mechanism to amplify other changes. However, there are no detailed records of sea-ice extent around the AP against which we can compare other proxy records and so it is difficult to assess the true role of sea-ice in AP palaeoclimate change. Elsewhere, the latest ice core evidence from Dome C shows that maximum sea ice extent is closely tied to Antarctic temperature on multi-millennial timescales, but less so on shorter timescales (Wolff *et al.* 2006). Studies of coastal ice core MSA records also provide a local to regional scale proxy for sea ice. The ice core MSA records show strong variability at inter-annual to decadal scales, however the factors dominating this variability clearly differ from region to region around Antarctica.This hints at a more complex relationship between Antarctic sea-ice and climate (Abram *et al*., 2007).

**Other mechanisms**

In the debate on the RRR warming, numerous mechanisms have been proposed including several sub-decadal modes of climate change. We are not able to evaluate these in the palaeoclimate record because the resolution of the record is insufficient, or because the mechanisms may leave no clear record in the proxy record of palaeoclimate change. For example, we do not consider here the Southern Hemisphere Annular Mode (SAM) - sometimes referred to as the Antarctic Oscillation (AAO) - or time-dependent variations in the upper-air long wave (Rossby wave) pattern around the Antarctic, which may cause circulation changes (Björck *et al*., 1996b). Changes in the SAM manifest as shifts in position or intensity of the westerlies (e.g. Marshall *et al*., 2006) but there is no clear understanding of any centennial or longer changes in this mode of variability.

**Thresholds and feedbacks**

A difficult but crucial question relates to the question of thresholds. A simple comparison of the generally smooth forcing mechanisms (Fig. 3) and the sometimes abrupt proxy records (Fig. 2) makes it clear that the system must be non-linear. Forcing mechanisms, probably acting in combination are likely to pass thresholds which can lead to abrupt changes. Identifying these thresholds is a major challenge across palaeoclimate science, not just in Antarctica.

There may also be positive or negative feedbacks operating. One example of this might be the effect of warming on snowfall. As warming continues in the western AP it may increase snowfall, which (coupled with increased Westerlies pushing the pack against the coast) can cause an extended period of sea ice coverage because of thick snow cover on the pack ice. This will mitigate against continued warmth. The snowfall effect is discussed by Massom *et al.* (2006) for the warm summer of 2001-2002, and for the palaeo record in Domack (2002).

## 4. DISCUSSION: LINKING FORCING MECHANISMS TO AP CLIMATE CHANGE DURING THE HOLOCENE

In this section we now attempt to link the forcing mechanisms listed above to each of the major periods of Holocene climate change on the AP. This also draws on palaeoenvironmental records elsewhere in Antarctica and in South America to try and understand what was happening in the high latitudes of the Southern Hemisphere. This approach is necessarily broad, but aims to identify what is known and where there are gaps and inconsistencies in our knowledge. Following this we identify some of the major challenges facing palaeoclimate researchers trying to understand Holocene change on the AP.

**Mechanisms for the early Holocene optimum (11-9.5 cal ka BP)**

The early Holocene climate optimum is expressed in Antarctic ice cores but the mechanisms behind it are not yet fully understood (Masson-Delmotte *et al*., 2004). It has been speculated that at the end of Northern Hemisphere deglaciation, reduced North Atlantic Deep Water (NADW) formation could have resulted in warmer conditions over Antarctica (Blunier *et al*., 1997; 1998). In effect, shutdown of the NADW meant that warm ocean water was no longer drawn away from high southern latitudes (Broecker, 1998; Blunier *et al*., 1998). If this theory is correct then the early Holocene optimum in Antarctica could have been a result of the 'switching off' of the thermohaline circulation during the Northern Hemisphere deglaciation, thereby restricting the removal of heat from the high southern latitudes (Broecker, 1998). The end of the early Holocene optimum in Antarctica could then have been brought about by the 'switching on' of the thermohaline circulation following the end of Northern Hemisphere deglaciation thereby removing heat from high southern latitudes (Broecker, 1998). However, there is no direct evidence in the AP proxy records for this mechanism driving early Holocene warmth and the impact of thermohaline change remains speculative. We cannot rule out indirect effects of thermohaline circulation change on other forcing such as the Southern Westerlies.

Annual solar insolation values were at their highest in the early Holocene, so the evidence of warmth could be related to increased radiation inputs. However, solar insolation declined gradually whereas the early Holocene warm period ended fairly abruptly. Thus, some other mechanism was likely involved, or there are non-linear thresholds in the AP climate system. Evidence that the Southern Westerlies were involved in the early Holocene climatic optimum can be seen in South American proxy records of rainfall associated with the southward transport of moist Pacific air and a poleward displacement of the Southern Westerlies (e.g., Lamy *et al*., 2002). From this, McCulloch and Davies (2001) inferred from pollen data in the Magellan Strait region of southernmost South America that the westerlies moved south between 11.4 and 9.5 cal ka BP (Fig. 3E). Similarly, Mayr *et al*. (2007) suggest from pollen evidence in Patagonia that the westerlies were located further south prior to 9.2 cal ka BP. Both of these studies show northwards migration of the westerlies consistent with the end of the early Holocene optimum and which would have reduced the influence of warm, moist air from the west side of the AP.  This pattern is also seen in SST data from the southeast Pacific (ODP Site 1233) (Kaiser *et al*., 2005). So early Holocene warmth may have been caused by high insolation and a polewards displacement of the westerlies.

This still leaves the issue of apparent cold conditions in the Palmer Deep at this time, which is difficult to reconcile with early Holocene warmth. One possible explanation is that they are in fact a local response to deglaciation with increased glacial meltwater causing more persistent sea ice and perhaps standstill or even readvance of grounded glacial ice around the Palmer Deep Basin. A further possibility is that the Palmer Deep proxies in this period may record discharge of ice from glaciers following collapse of buttressing ice shelves through mechanisms seen on the AP today (Rott *et al*., 2002; De Angelis & Skvarca 2003; Scambos *et al*., 2004; Rignot *et al*., 2004). Sediment proxy records of such relatively cold conditions immediately adjacent to a retreating margin have been described for the East Antarctic shelf (Leventer *et al*., 2006), and were interpreted to reflect conditions in calving bay reentrants along the margin. Calving bays were thought to be present during deglaciation of the inner shelf of the western AP (Domack *et al*., 2006) and so perhaps the deglaciation of Palmer Deep itself was accompanied by a calving bay. However, the varved sequence (~ 13.2-11.5 cal ka BP) that might represent the record of a calving bay reentrant is distinct from, and pre-dates, the record of colder conditions (~ 11.5-9 cal ka BP) in Palmer Deep and so the calving bay model may not explain the cold conditions. Alternatively, it could be argued that the southerly position of the westerlies would have increased precipitation on the west side of the AP and so may be related to the apparent evidence of glacier advance seen in the Palmer Deep.

*After the optimum I (9500 -7945 cal yr BP)*

The retreat of the currently extant George VI Ice Shelf immediately after the atmospheric (ice core) optimum suggests that that atmospheric temperatures had possibly reached the highest point so far experienced in the Holocene and had eroded the surface of the ice shelf over a long period of time and predisposed it to collapse (Bentley *et al*., 2005b; Smith *et al*., 2007). The retreat was coincident with an inferred influx of relatively warm water onto the AP continental shelf identified in George VI Sound by the presence of foraminifera species characteristic of a high-productivity, upwelling water mass (Bentley *et al*., 2005b; Smith *et al*., 2007). Thus George VI Ice Shelf was probably melted from above and below.

Warm surface waters were also recorded at the Palmer Deep from 9000 to 6700 cal yr BP (Domack, 2002; Leventer *et al*., 2002) and, whilst the record does not necessarily imply a wholesale shift of water masses onto the shelf it does indicate some degree of surface layer stability and heat gain through the summer season. This explanation would account for the foraminiferal oxygen isotope record, which indicates a lack of CDW at this time (see also Ishman and Sperling, 2002). Nevertheless, Shevenell and Kennett's (2002) marine isotope study at Palmer Deep argues for a sustained presence of CDW between 9000 and 3600 cal yr BP with warmer regional atmospheric and sea-surface temperature, decreased sea-ice cover and increased primary production. This conclusion is consistent with other proxy evidence (e.g., diatoms) from the Palmer Deep, which indicates the presence of 'warm water conditions' between 9000 and 6700 cal yr BP (Leventer *et al*., 2002; Taylor and Sjunneskog, 2002). However, this interpretation is at odds with the foraminiferal assemblage study of Ishman and Sperling (2002), which suggests that CDW was absent on the western AP shelf during this time interval and was dominated instead by the production of cold saline shelf water in the *absence* of CDW. This is a key difference in interpretation of the Palmer Deep record because much of our understanding of the role of CDW in the early to mid-Holocene on the mid-shelf hinges on how this is resolved.

The role of the Westerlies during this period is not yet fully understood. However, as already noted, pollen records in southern Patagonia suggest they moved north by 9.5 - 9.2 cal ka BP (Fig. 3E) (McCulloch and Davies, 2001; Mayr *et al*., 2007) but they may not have reached their modern configuration at this time. For example, Jenny *et al*. (2003) used lake-level records from central Chile (34 °S) to infer that the Westerlies were located further south than present during the early to mid-Holocene. They suggested that there was a two step increase in precipitation in Chile (linked to southward movement of the Westerlies) at 8000 and 6000 cal yr BP. Therefore it may be that the period between the early Holocene warm period and the MHH saw an increasing influence of the westerlies, possibly even with stepped change (Fig. 3e), but this is certainly not well-constrained. Other forcing mechanisms for climate change in this period have been suggested. Specifically, the marine core evidence of cooling in the South Atlantic following 9 cal ka BP has been attributed to expansion of the winter sea-ice field (Bianchi and Gersonde, 2004; Nielsen *et al*., 2004).

Therefore this period shows a complex pattern of change with relative warmth on the western side of the Peninsula (Palmer Deep and George VI Ice Shelf collapse), but cooling in the South Atlantic. The Southern Westerlies may have lingered in the latitudes of the AP and allowed continued warmth, but at the same time the sea-ice may have been expanding in the South Atlantic. The position of the Southern Westerlies is sensitive to the position of the sea-ice edge so these two mechanisms are difficult to reconcile unless the sea-ice was showing very different behaviour in the western AP and South Atlantic sectors. Until this issue is resolved, it seems that the western AP and eastern AP may have been subject to very different influences during this period.

*After the optimum II (7945-4500 cal yr BP)*

Atmospheric and/or oceanographic cooling continued and by 7945 cal. yr BP the George VI Ice Shelf had reformed in response to cooler climatic conditions, implying that either the influence of CDW was reduced in George VI Sound from approximately 7.5 cal ka BP, or that the ice shelf responded to an atmospheric cooling that allowed the ice shelf to overcome the effects of basal melting (Smith *et al*., 2007). It seems likely that the Holocene stability of George VI ice shelf depended on a fine balance between atmospheric temperatures and intrusion of warmer water masses. Some Antarctic ice core data (e.g., Taylor Dome) show a temperature minimum (cold event) at ~ 8000 cal yr BP, followed by a secondary warm event around 6000 cal yr BP (Stager and Mayewski, 1997; Steig *et al.*, 1998; Masson *et al.*, 2000; Masson-Delmotte *et al.*, 2004).

In summary, this period shows a more consistent picture than 9500-7945 cal yr BP – cooling, or continuing cool conditions, were prevalent across many records on the western AP and in the South Atlantic. The model of Renssen *et al*. (2005) shows that October (spring) and July (winter) temperatures gradually declined from 9 cal ka BP (Fig. 3d). Thus, progressive cooling in this part of the Holocene may have been driven largely by solar insolation change during winter and spring, and some South American proxy records of the Westerlies suggest they had reached a near-modern configuration by 6 cal ka BP, with a commensurate reduction in influence on the western AP.

**Mechanisms for the mid-Holocene optimum (4.5-2.8 cal ka BP)**

Orbital calculations (Berger and Loutre, 1991) show *summer* insolation at the latitude of the AP (65°S) has been increasing over much of the Holocene (from a minimum at ca 10 cal ka BP to a Late Holocene maximum and a very small decrease to the present day) (Fig. 3a). At face value this increase in insolation over the Holocene could suggest that the MHH may have been initiated through increases in solar insolation. This is supported by the modelling of Renssen *et al*. (2005) where January (summer) temperatures reach an optimum at ~4-3 cal ka BP (Fig. 3d).

An alternative explanation for the MHH is that a poleward displacement of the Southern Westerlies brought warm, moist air to the west side of the AP leading to higher temperatures and precipitation. A southwards shift in the Southern Westerlies might have led to an intensification of the ACC, particularly through the Drake Passage, and this could be sufficient to drive CDW onto the western shelf of the AP. If there was sufficient mixing with surface waters then this could potentially reduce sea-ice extent driving further warming in the western AP. A southward shift of both the Southern Westerlies and the ACC during the middle Holocene is supported by ice core (Thompson et al., 1998) and lake records (Cross *et al.*, 2000) from the South American Altiplano, which indicate increased aridity at these latitudes, and could relate to a more intense Hadley cell. South of the Altiplano, a shift of the wind belt to the south may have involved an increase in moisture from subtropical sources and may explain the synchronous warm events experienced in South America c. 3330 to 2230 $^{14}$C yr BP (~3550 to 2250 cal yr BP) (Clapperton and Sugden, 1988). There is also evidence for decreased west Antarctic sea ice during this period (Stager and Mayewski, 1997). Björck *et al*. (1993, 1996b) attribute the MHH in the AP to the presence of warmer and more humid air masses resulting from this decrease in sea ice around Antarctica. A southwards shift in Southern Westerlies would also explain the increase in South American pollen seen in mid-Holocene lake sediments in the northern AP (Björck *et al*., 1993). However, the SST records in the SE Pacific do not seem to support the notion of a southwards shift of the Westerlies at this time (Fig 2d).

At this time the Palmer Deep record shows that the long Holocene climate optimum there (9.07-3.36 cal ka BP) was coming to an end with a slow cooling, and reduced sediment accumulation through the MHH (Domack, 2002). This suggests that the warming seen in terrestrial records was *not* accompanied by a discrete period of oceanographic warming, at least at this location.

In summary, either or both of solar insolation and the Southern Westerlies may have played a role in the MHH. A fundamental issue here is that although there is broad agreement between the terrestrial records (lake sediments and moss banks), selected ice cores (e.g. Plateau Remote) and selected marine records (Lallemand Fjord, Prince Gustav Channel Ice Shelf, Larsen-A Ice Shelf ) the Palmer Deep marine record does not show a contemporaneous MHH signal. This is one of the most puzzling aspects of understanding Holocene palaeoclimate change on the AP – why does a prominent thermal optimum on land not show up in one of the most widely used palaeo-oceanographic records? One possible explanation is that the proxies in these environments are responding to environmental conditions expressed at different times of the year. In other words, the Palmer Deep record reflects differences in the forcing during different seasons (Fig. 3a, d). For example, if the Palmer Deep record reflects the springtime phytoplankton bloom then this would be consistent with insolation and temperatures in spring months, which reached a maximum several kyr prior to the insolation and temperatures in summer (Renssen *et al*., 2005). This is consistent with the warming in the Palmer Deep record being a few kyr earlier than in the terrestrial record, whereas productivity in onshore lakes may have been responding to summer insolation-driven loss of lake ice and catchment snow melt. Moreover, the prominence of the event in the terrestrial records might also be partly explained by positive albedo feedbacks that would have enhanced snow melt in lake catchments. A further alternative explanation is that because the Palmer Deep is on the mid-shelf it records different environmental changes to those seen in the inner shelf or onshore. Clearly, much further work is required to understand the spatial pattern and mechanisms that drove the MHH.

*After the MHH: Neoglacial and the Medieval Warm Period*

The onset of the Neoglacial (3.36 cal ka BP) in the Palmer Deep record predates the record of cooling from lakes, but is consistently recognized in all Palmer Deep paleoenvironmental proxy data (Domack *et al.*, 2001; Domack, 2002) including a decrease in Mass Accumulation Rate (MAR) and increase in coarse-fraction IRD (Domack, 2002). Shevenell and Kennett (2002) suggest that this may reflect a general increase in the presence of shelf water (replacing CDW) and westerly wind strength between ~ 3600 and 50 cal yr BP resulting in a general cooling. They suggest that predominantly offshore winds could push the southern boundary of the ACC further away from the western AP continental shelf, thereby depressing the volume of CDW in the Palmer Deep (cf. Hoffman *et al*., 1996; Smith *et al*., 1999). It is during this interval that smaller-scale mechanisms become resolved. For example, there is wide agreement that the intensity of CDW flow and its movement across the shelf fluctuated many times in the last 3.7 kyr (Shevenell and Kennett, 2002; Ishman and Sperling, 2002). This resulted in the alternating periods of more intense (seasonally persistent) sea ice and open water seen in the diatom record (Taylor and Sjunneskog, 2002; Sjunneskog and Taylor, 2002). The importance of CDW as a mechanism for the neoglacial period is discussed in Shevenell and Kennett (2002) who use benthic foraminiferal records to document consistent and rapid alternations in bottom water temperatures on the shelf with amplitudes of 1.0° to 1.5°C, and suggest that these changes reflect atmospheric forcing via westerly wind strength on the axial flow of the ACC.

Thus the Southern Westerlies may be implicated in fluctuations of oceanographic conditions on the western AP shelf following the Neoglacial cooling. Continuing warmth on land (i.e., the ongoing MHH) may be attributable to the relatively high summer insolation and the westerlies. The 'flickering' variability in oceanographic conditions is also consistent with the highly variable ENSO signal that reached a peak by 1200 cal yr BP (Moy *et al*., 2002).

The MWP has not been unequivocally recorded in Antarctic records, but despite this a few studies have looked at potential driving mechanisms. For example, Goosse *et al*. (2004) suggest that changes, and in particular warm surface anomalies in the North Atlantic, can be transmitted via the thermohaline circulation to the Southern Ocean, with a lag of ~150 years. The heat is then released around Antarctica by large-scale upwelling. Broecker (2001) has also suggested that a global MWP was driven by changes in North Atlantic thermohaline conditions.

There is no obvious change in AP solar insolation in the Late Holocene that might have caused a MWP. Indeed, such a short-lived event is unlikely to be caused by insolation cycles, unless very strong non-linear feedbacks are acting. We are not aware of any proxy record of Southern Westerly movements in the MWP, and so it is difficult to evaluate the potential influence of this mechanism. Concentrations of $CO_2$ reached a pre-industrial peak at ~ 1000 cal yr BP (Fig. 3b), and showed a downturn in the few centuries following. It is possible that this partly drove the warmth of the MWP and the subsequent downturn to the LIA. However, the concentration changes involved are relatively small and modelled temperatures do not show a strong response to the change in $CO_2$ (Renssen *et al*., 2005). In summary, we know very little about the MWP, and there is no consensus of evidence for such an event in the AP.

*Little Ice Age*

The Little Ice Age has been recognized in Palmer Deep from 0.7 to 0.15 cal ka BP and in several glacier advances. Over this period the pelagic and hemipelagic record of more persistent sea ice, colder sea-surface and bottom-water conditions in Palmer Deep do indeed correlate with local glacial advances and ice core records (Domack *et al*., 1995; Shevenell *et al*., 1996; Leventer *et al*., 1996, 2002; Shevenell and Kennett, 2002; Taylor and Sjunneskog, 2002; Sjunneskog and Taylor, 2002; Warner and Domack, 2002). However, the LIA has not been widely noted in lake core records. As noted above its onset might be related to a downturn in greenhouse gas concentrations. Moreover, the modelled temperatures in all seasons show minima in the last few centuries (Fig. 3d) (Renssen *et al*., 2005) suggesting that the LIA may have been driven by the decline of solar insolation through the Holocene. Moreover, other studies have suggested that the association of the LIA with a minimum in solar irradiance (known as the Maunder minimum) may indicate a dominant role of solar forcing for the LIA (e.g. van Geel *et al*., 1999).

**Recent Rapid Regional warming**

Above, we have evaluated what is known about the periods of Holocene warmth that may be potential past analogues for the processes accompanying the RRR warming now being experienced in the AP. These are the early Holocene warm period, the mid-Holocene warm period, and the Medieval Warm Period. Here, we discuss mechanisms for the RRR in the context of the understanding gained so far from investigation of Holocene warm periods.

Available records show that the RRR warming of the western AP has apparently been very abrupt yet has occurred during gradually decreasing solar insolation, so a solar forcing mechanism can be ruled out. Previous explanations have included both atmospheric and oceanic forcing mechanisms.

In terms of atmospheric mechanisms there have been numerous investigations of recent meteorological data to try and understand the reasons for the RRR warming. What is now clear is that anthropogenic greenhouse gases have caused at least part of the warming in the AP (e.g. Marshall *et al*., 2006), and that this has caused increased westerly winds in the northern AP. Shindell and Schmidt (2004) have described a southwards movement of the westerlies in recent decades. Marshall *et al*. (2006) have shown how the increased westerlies have created a strong foehn wind effect in the north-eastern AP, which is directly implicated in the warming that led to the break-up of the Larsen Ice Shelf. Moreover, the strong sea ice feedback in the western AP (King, 1994) helps account for the exceptional magnitude of the RRR temperature change in the western AP.

Oceanic mechanisms have also been suggested (e.g., recent intrusions of CDW onto the continental shelf) along with teleconnections to ENSO (Trenberth and Caron, 2000), but so far these mechanisms are less well-constrained, partly because of poorer coverage of oceanographic measurements. On the western side of the AP, Meredith and King (2005) have demonstrated recent warming of surface layers of the ocean and strong upper-layer salinification. This is similar to some of the changes described from proxy records for the early Holocene (e.g. Leventer *et al*., 2002; Taylor and Sjunneskog, 2002).

*Common features of palaeoclimate records and RRR*

It is worth summarising some of the features that we suggest are common to both palaeo-records and contemporary records of climate change. Firstly, there appears to be a strong contrast between the west and east sides of the AP. Past and current warming events seem to be more strongly developed on the west side of the AP than to the east. We know that there is a strong contemporary environmental contrast between the two sides of the AP (e.g. Reynolds, 1981; Domack *et al*., 2003a), and Nielsen *et al*. (2004) discuss the possibility of a Holocene palaeoclimatic divide through the Drake Passage, with climate patterns in the area to the west of this divide (western AP, southernmost South America) being in phase with one another, but out of phase with patterns to the east of the divide (South Atlantic, NW Weddell Sea). Further, sharp contrasts in deglacial ice sheet and ice shelf behaviour between east and west have been discussed specifically by several authors (Evans *et al*., 2005; Hodgson *et al*., 2006; Sugden *et al*., 2006). Secondly, movement of the Southern Westerlies may have forced a number of the prominent palaeoclimate events, and is implicated in the RRR warming (Shindell and Schmidt, 2004; Marshall *et al*., 2006). Specifically, the early Holocene warmth and mid-Holocene periods seem to have coincided with polewards shifts of the Southern Westerlies warming. Thirdly, there seems to be a growing body of evidence that teleconnections between ENSO and the (western) AP may have forced some palaeoclimate changes in the AP, and can explain some of the changes associated with the RRR.

## 5. FUTURE RESEARCH GOALS

It is clear from this synthesis that there are still substantial gaps in our knowledge of past environmental changes in the AP region. As one of the fastest warming regions on Earth it merits further study to enable researchers to unravel the causes and consequences of past and ongoing climate change. We suggest the following research priorities:

*Ice coring*

There are no long ice cores from the AP and there is a need to find a site where it is possible to retrieve ice from the glacial-interglacial transition. This has proved difficult because of the shallow and complex topography along the spine of the AP, and high heat flow in the northern parts. This has meant that cores have been either very short (few centuries) or have been drilled further south (Siple Dome) or east (Berkner Island), and are not fully representative of AP conditions. However, recent drilling in 2008 at James Ross Island looks likely to have captured the full Holocene (R. Mulvaney, *pers. comm*., 2008). We also suggest here that shorter cores, stretching to either the early Holocene, or perhaps even only to the mid-Holocene would yield enormously important information on the spatial pattern and mechanisms of several of the changes we have discussed above. In particular, it would be helpful to have a regional record of atmospheric temperature and precipitation during the mid-Holocene. Mosley-Thompson and Thompson (2003) outline some of the challenges to such coring, and further questions that AP ice cores might address.

*Marine and terrestrial records*

There is a further need to extend the network of marine and terrestrial (lake) geological records so that the periods when the terrestrial and marine records appear to show significantly different behaviour - or are even 'decoupled' - such as during the mid Holocene warm period, can be examined more closely. Spatial patterns are also important and some recently-sampled transects by various groups will help address issues of spatial contrasts in palaeoclimate.

Radiocarbon dating is of fundamental importance when comparing marine and terrestrial records in the Antarctic. The well-known issues of marine reservoir corrections and incorporation of 'old' reworked carbon into marine sediment make accurate dating of marine sediment a real challenge. Whilst progress has been made, and some records such as Palmer Deep (Domack, 2002) have yielded apparently robust chronologies there is a continued need for rigorous dating strategies, incorporating where appropriate use of carbonate and bulk paired dates, dating specific organic geochemical fractions in the sediments (e.g., Ingalls *et al*., 2004; Ohkouchi and Eglinton, 2008), step-combustion (e.g., Schrum *et al*., 2006; Rosenheim et al., 2008), and alternative techniques such as palaeomagnetic intensity dating (e.g., Brachfeld *et al*., 2003; Willmott *et al*., 2006) .

There is potential for greater study of the most recent palaeoclimate changes in the AP, namely the MWP, LIA, and proxy record of the RRR warming. Since these provide the most recent context for RRR then improved understanding of the proxy record in these intervals will give us clues as to how atmospheric (e.g. ENSO/Southern Westerlies) and oceanic (e.g. changes in ACC/CDW) forcing, plus sea-ice feedbacks might interact with or drive this change in the near future.

*Role of sea ice.*

Sea-ice is known to be a key factor in the modern climate of the AP. Indeed, the west coast of the AP is the only place in Antarctica where there is a clear correlation between sea ice extent and coastal temperatures (King, 1994; King and Turner, 1997; Jacobs and Comiso, 1997). In order to determine the role of sea ice in Holocene change we require more information on the past distribution of sea ice, particularly west (upwind) of the AP. The analyses for sea ice proxies exist and have been proven (e.g.

Crosta *et al*., 1998; Burckle, and Mortlock, 1998; Leventer, 1998; Gersonde and Zielinski, 2000; Gersonde *et al*., 2005), but finding appropriate, sufficiently high-resolution sediment cores that can be reliably dated remains a major challenge around the AP.

*Circumpolar Deep Water*

A role for CDW has been hypothesised for some of the palaeoenvironmental changes we have discussed in this paper. However, for some of the suggested mechanisms there is disagreement on the sign of change. For example, there have been suggestions that CDW may be brought onto the shelf by both decreased (Shevenell and Kennett 2002) and increased (Smith *et al*., 1999; Smith and Klinck 2002) westerly wind strength. A further challenge for oceanographic modellers is to fully understand linkages to the Southeast Pacific, such as changes in the ACC or movements of CDW.

Previous studies have used particular faunal assemblages to infer presence/absence of CDW (e.g., Ishman and Sperling, 2002) but this is laborious and has potential difficulties in interpretation. Moreover, attempts to infer its past presence from (co-)isotopic analysis of carbonate organisms (e.g., Shevenell and Kennett, 2002; Smith *et al*., 2007) has been hampered by a lack of contemporary datasets on the isotopic signature of CDW and other water masses around the AP (e.g., Mackensen *et al*., 2001), and by poor preservation of calcareous microfossils in the marine sediments. It would therefore be helpful if we could develop a better contemporary dataset of isotopic composition of modern faunal assemblages in different oceanographic regimes (e.g., by the deployment of moorings on the AP shelf which record hydrographic variability, in particular CDW incursions, and which are fitted with sediment traps for the capture of tests of planktonic organisms) and even to find a new proxy for CDW that could be detected in sediment or geochemical analyses.

*Models*

Better two-way linkages between modellers and ‘field’ palaeoclimatologists will greatly aid understanding of the past patterns of climate change in the AP region. In turn, this will help to inform a new generation of climate models that are regionally sensitive, and to stimulate field programmes to collect data to constrain such models. Some such models already exist: for example in this paper we have used the model simulations of Renssen *et al*. (2005) to explain some of the contrasts between marine and terrestrial records in the mid-Holocene.

## 6. CONCLUSIONS

1. Two warm events are well recorded in the Holocene palaeoclimate record of the AP, namely the early Holocene warm period, and the Mid Holocene Hypsithermal (MHH). Two are less well-

recorded in proxy records – the Medieval Warm Period (MWP) and the Recent Rapid Regional warming.

2. We have suggested that the early Holocene warm periods and MHH might be explained by relatively abrupt shifts in position of the Southern Westerlies, superimposed on slower solar insolation changes.
3. Evidence for a MWP (and subsequent Little Ice Age) is patchy on the AP. If they did occur then both of greenhouse gas and solar insolation changes provide possible mechanisms.
4. At least some of the marine and terrestrial records in the AP appear to be significantly different, or even 'decoupled', during the MHH. The MHH manifests as a prominent warming in terrestrial records but in the Palmer Deep core – the longest and best-dated marine record – the warming is much earlier and a MHH is not evident.
5. We have suggested that some of the differences in marine-terrestrial behaviour might be explained by contrasts in the seasonality to which these records are responding: spring solar insolation (driving marine productivity) peaked earlier in the Holocene than the summer insolation (driving lake productivity after ice melt).
6. Circumpolar Deep Water probably played a key role, particularly in driving change in the western AP, but more work is needed to understand how we might track movements of this water mass in proxy records. It has been implicated in several of the prominent changes through the Holocene, but there are still differences in how its presence/absence should be interpreted from proxy records.
7. Further work is required in several areas, notably understanding seasonality and contrasts between marine and terrestrial records, east-west contrasts in palaeoclimate, the history of CDW, a long onshore high resolution record of the Holocene (probably ice core), and the role of sea ice in driving or modulating palaeoclimate change. Further, an increased focus on the MWP, Little Ice Age and RRR intervals in existing and new palaeorecords would help our understanding of these events, and in particular assess the long-term context and significance of the Recent Rapid Regional Warming of the AP.

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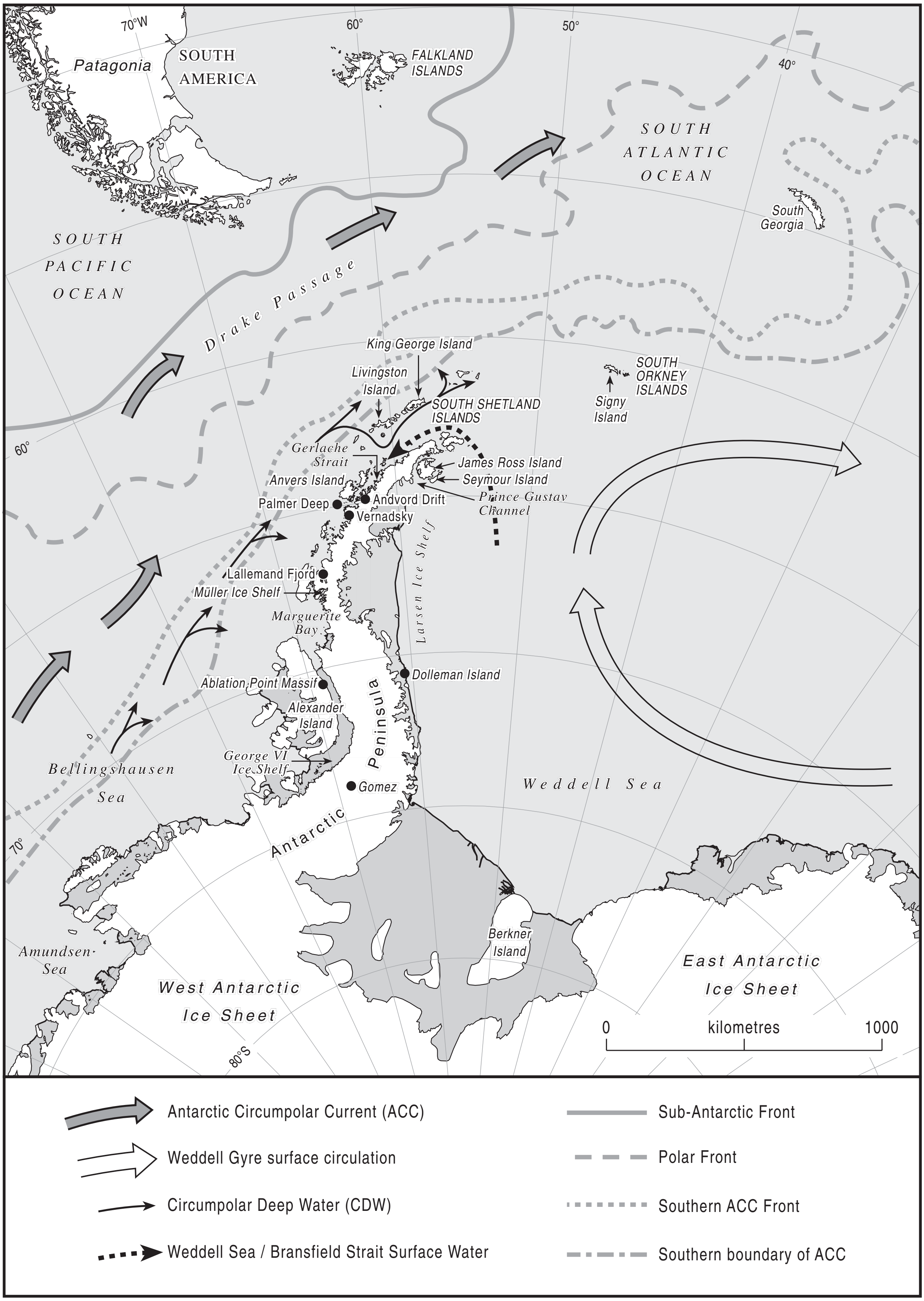
## List of Figures

**Figure 1.** Setting of the Antarctic Peninsula and southernmost South America showing sites discussed in text. Ocean circulation and frontal boundaries are from Hernández-Molina *et al*. (2006), Orsi *et al*. (1995) and Naveira Garabato *et al*. (2002).

**Figure 2**. Proxy records of climate change since ~ 14 cal ka BP. (A) Ice core synthesis for the circum-Antarctic (Masson-Delmotte *et al*., 2004). Data are potted as anomalies from the Holocene mean; (B) Magnetic susceptibility of the marine core from the Palmer Deep and its interpretation (Domack, 2002); (C) Concentrations of *Alaskozetes antarcticus*, an oribatid mite, in lake sediments from Heywood Lake, Signy Island, South Orkney Islands (Hodgson and Convey, 2005). Higher mite concentrations are interpreted as indicating warmer and wetter conditions; (D) Total Organic Carbon in a Lallemand Fjord core (Shevenell *et al*., 1996) and Sea-Surface Temperatures in the Peru-Chile Current (PCC) (Lamy *et al*., 2002; Kaiser *et al*., 2005). Note that the two SST records come from different cores but in closely similar locations off the coast of south-central Chile. (E) Multi-archive compilation of mid-Holocene hypsithermal (MHH) records in selected ice, lake, marine and glacial records from the Antarctic Peninsula (redrawn from Hodgson et al., 2004). Shaded bands show where relative warmth has been interpreted. References: [1]Ingólfsson et al. (1998), [2]Ciais et al. (1994), [3]Jones et al. (2000), [4]Björck et al. (1993), [5]Björck et al. (1996b), [6]Domack and McClennen (1996), [7]Pudsey and Evans (2001).

**Figure 3.** Potential forcing mechanisms for palaeoclimate change in the AP since ~ 12 cal ka BP. (A) Solar insolation. Data are plotted as deviations from present-day means, calculated using *AnalySeries* (Paillard *et al*., 1996) with the Laskar *et al*. (2004) solution; (B) Greenhouse gases (data from Raynaud *et al*., 2000 but smoothed by Renssen *et al*., 2005); (C) strength of ENSO (Moy *et al*., 2002). The data show number of ENSO events per 100 years with the threshold for modern ENSO-type behaviour shown as a horizontal dashed line; (D) Modelled temperatures for the Holocene (Renssen *et al*., 2005). See text for discussion of model assumptions; (E) Schematic plot showing latitudinal shifts in westerlies (drawn using proxy data from: McCulloch and Davies, 2001; Mayr *et al*., 2007; Jenny *et al*., 2003; Lamy *et al*., 2001; 2002; Kaiser *et al*., 2005).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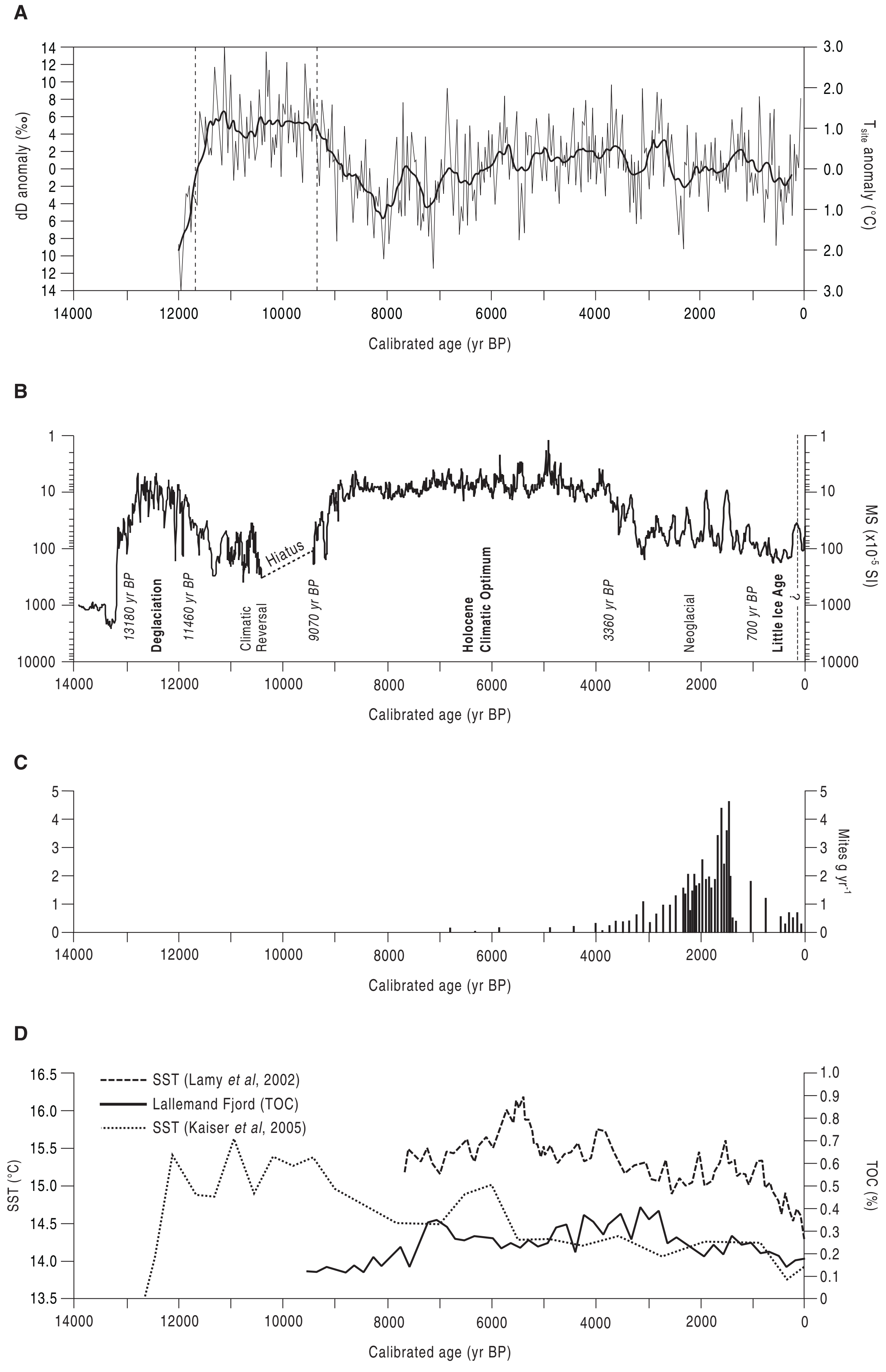
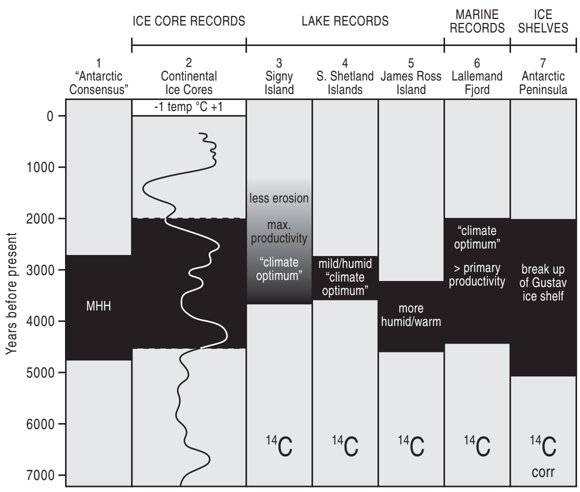
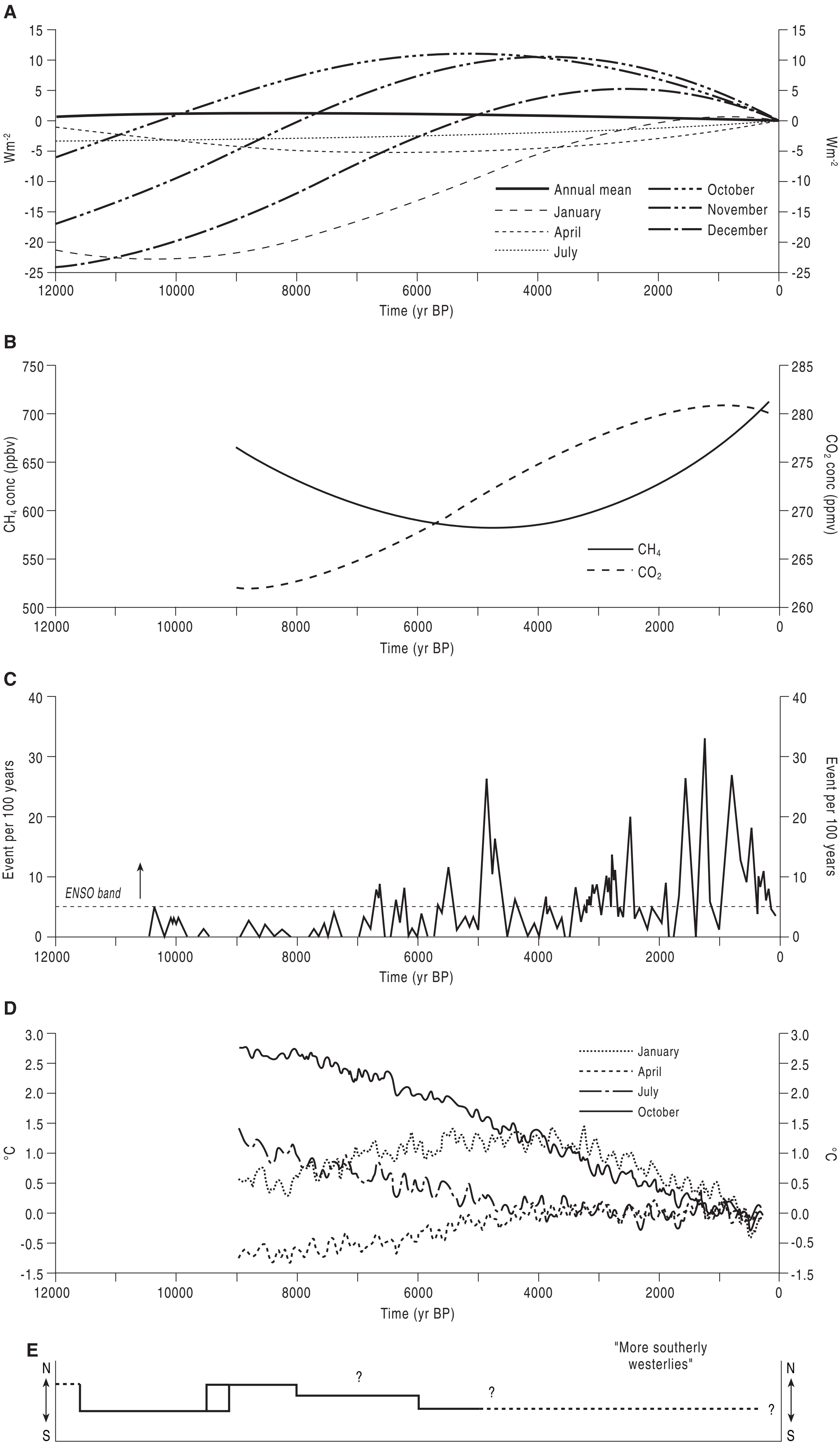
South Georgia
SOUTH PACIFIC OCEAN
Drake Passage
King George Island
Livingston Island
SOUTH SHETLAND ISLANDS
SOUTH ORKNEY ISLANDS
Signy Island
Gerlache Strait
Anvers Island
James Ross Island
Seymour Island
Prince Gustav Channel
Palmer Deep
Andvord Drift
Vernadsky
Larsen Ice Shelf
Lallemand Fjord
Müller Ice Shelf
Marguerite Bay
Dolleman Island
Ablation Point Massif
Alexander Island
George VI Ice Shelf
Gomez
Antarctic Peninsula
Bellingshausen Sea
Weddell Sea
Amundsen Sea
Berkner Island
West Antarctic Ice Sheet
East Antarctic Ice Sheet
0
kilometres
1000
Antarctic Circumpolar Current (ACC)
Weddell Gyre surface circulation
Circumpolar Deep Water (CDW)
Weddell Sea / Bransfield Strait Surface Water
Sub-Antarctic Front
Polar Front
Southern ACC Front
Southern boundary of ACC


A
dD anomaly (‰)
Tsite anomaly (°C)
Calibrated age (yr BP)
B
Hiatus
13180 yr BP
Deglaciation
11460 yr BP
Climatic Reversal
9070 yr BP
Holocene Climatic Optimum
3360 yr BP
Neoglacial
700 yr BP
Little Ice Age
?
MS (x10-5 SI)
Calibrated age (yr BP)
C
Mites g yr-1
Calibrated age (yr BP)
D
SST (Lamy et al, 2002)
Lallemand Fjord (TOC)
SST (Kaiser et al, 2005)
SST (°C)
TOC (%)
Calibrated age (yr BP)


ICE CORE RECORDS
LAKE RECORDS
MARINE RECORDS
ICE SHELVES
1 "Antarctic Consensus"
2 Continental Ice Cores
3 Signy Island
4 S. Shetland Islands
5 James Ross Island
6 Lallemand Fjord
7 Antarctic Peninsula
-1 temp °C +1
Years before present
0
1000
2000
3000
4000
5000
6000
7000
MHH
less erosion
max. productivity
"climate optimum"
mild/humid "climate optimum"
more humid/warm
"climate optimum"
> primary productivity
break up of Gustav ice shelf
14C
corr


A
Wm-2
Annual mean
January
April
July
October
November
December
Time (yr BP)
B
CH4 conc (ppbv)
CO2 conc (ppmv)
CH4
CO2
Time (yr BP)
C
Event per 100 years
ENSO band
Time (yr BP)
D
°C
January
April
July
October
Time (yr BP)
E
"More southerly westerlies"
N
S
?